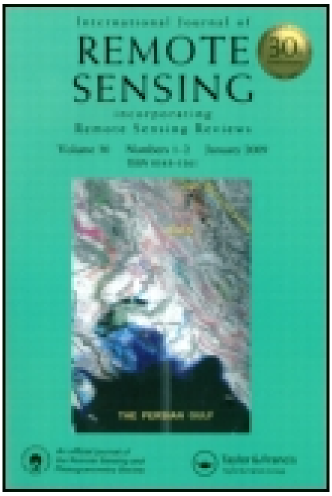

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Cloud optical parameters from airborne observation of diffuse solar radiation accomplished in USA and USSR in different geographical regions

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Cloud optical parameters (optical thickness, single scattering albedo, and ground albedo) are obtained from airborne experiments with NASA's Cloud Absorption Radiometer and analysed taking into account observational and processing uncertainties. The analytical approach of the inverse asymptotic formulas of the transfer theory, which uses observed values of solar diffuse radiance, is applied. The method is free from *a priori* restrictions and links put to desired parameters. The algorithms and first results of processing have been presented earlier. The first results, being the solution of the inverse problem, showed strong fluctuations in values, which required the regularization of the solution. The dependence of uncertainties of the result on viewing direction was revealed. Hence, here attention is focused on uncertainties of observation, angle function calculation, and processing approach, which is taken into account for result averaging, and the regularization procedure is described. Calculating the uncertainties of the processing approach is accomplished analytically using formulas for the retrieval of the optical parameters. The values of the desired parameters obtained in eight observational spectral channels – above, below, and within the cloud – at 16 levels are presented. The final results are compared to the optical parameters of extended cloud layers obtained earlier using a similar method of inverse asymptotic formulas from spectral data of Russian aircraft solar irradiance measurements in different regions, made in the 1970s/1980s at Leningrad (now Saint Petersburg) University in the USSR.

1. Introduction

In the past, the solution of inverse problems of planetary atmospheres in many cases involved model calculations with the subsequent comparison to the measured radiation characteristics and selection of optical model parameters for the best satisfying observed data for the atmosphere of Venus (Konovalov and Lukashevitch 1981; Maiorov et al. 2005) and for the Earth's atmosphere: (Rublev, Trotsenko, and Romanov 1997; Melnikova 1978). Often certain assumptions including *a priori* restrictions on desired parameters and links between optical parameters for different wavelengths are suggested, which prevents the realization of true values. One of the assumptions is conservative radiation scattering in the atmosphere in the shortwave ranges (King 1987; King, Radke, and Hobbs 1990; Rublev, Trotsenko, and Romanov 1997; Kokhanovsky et al. 2006; Rozanov and Kokhanovsky 2004) and the other is semi-infinite optical thickness of cloud (Yanovitskij 1972; Rozenberg et al. 1974). In some studies, simultaneous observations in

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two spectral channels were used (King 1987; King, Radke, and Hobbs 1990; Kokhanovsky et al. 2006).

The analytical method for retrieving cloud optical thickness from solar diffused radiance was proposed by King (1987), but accomplished only with conservative scattering assumption. The analytical approach based on inversion of the transfer theory asymptotic formulas for simultaneous retrieval of two cloud parameters – optical thickness and single scattering albedo (together with volume scattering and absorption coefficients) at every available wavelength – was derived and applied to airborne observational data by Melnikova (1991). Then, a similar method was modified for different observational geometry (satellite and ground-based observations of solar radiance and irradiance in cloudy atmospheres) (Melnikova and Mikhailov 1994, 2001; Melnikova, Domnin, et al. 2000; Melnikova and Nakajima 2000). Every type of observation has its advantages: satellite data provide global distribution and space averaging, ground experiments are cheap and available, airborne measurements allow study of inner cloud structure.

Here the method presented for airborne observation of diffused solar radiance at the cloud top, base, and within the cloud (Melnikova, Nikitin, and Gatebe 2009; Genya, Melnikova, and Gatebe 2010) is applied to data obtained with the Cloud Absorption Radiometer (CAR) instrument, NASA (Gatebe et al. 2003). A comparison of optical parameters, obtained from airborne radiative National Aeronautics and Space Administration (NASA) observations (that include shortwave diffused radiance) and older Russian ones (shortwave semispherical fluxes) in different geographical sites is shown.

2. Observational data

Experimental data obtained from NASA's Goddard Space Flight Center were obtained with a CAR (Gatebe et al. 2003; Román et al. 2011) in a cloudy atmosphere for above, below, and within the cloud layer in the framework of the programme, 'Southern African Regional Science Initiative 2000', on 13 September 2000 above the coast of South Africa at latitude 20.0–21.7° S and longitude 13.0–13.7° E. Flight altitude ranged from 354 to 1170 m, with levels above the cloud layer (800–1178 m), within (631–790 m), and below (343–400 m). The geometrical thickness of the cloud layer was 400 m (400–800 m).

The CAR instrument measures diffuse solar radiation in eight spectral channels: 340, 381, 472, 682, 870, 1035, 1219, and 1273 nm. The data contain geographical coordinates of the observation site, local time, solar zenith angle θ_0, azimuth angle relative to the Sun φ, altitude of the observation, and 182 values of radiance in reflectance units I_i in zenith viewing angles θ_i in ranges from −1 to 181° in 1° steps, as shown in Figure 1, for two spectral channels 0.34 and 0.682 μm. Measurements within the cloud were obtained too close to the top (altitude 799–631 m) and do not satisfy the diffuse domain condition, as shown in Figure 2, as per the maxima on the solar aureole.

The older Russian (USSR) airborne spectral data obtained from airborne experiments at Leningrad (Saint Petersburg) State University (Kondratyev et al. 1977; Grishechkin and Melnikova 1989; Grishechkin, Melnikova, and Shults 1989; Melnikova and Vasilyev 2004) in stratus clouds in different geographical regions were used earlier for optical parameter retrieval (Melnikova and Mikhailov 1994, 2001; Melnikova, Domnin, et al. 2000; Melnikova and Vasilyev 2004). These observations were accomplished within international research programmes: Global Atmospheric Tropical Experiment (GATE) programme in 1974 over the Atlantic Ocean off the coast of northwest Africa (17° N, 12 July 1974 and 4 August 1974) (Kondratyev et al. 1977); Complex Atmospheric Energetic Experiment (CAENEX) over the Black and Azov Seas (45° N, 10 April 1971

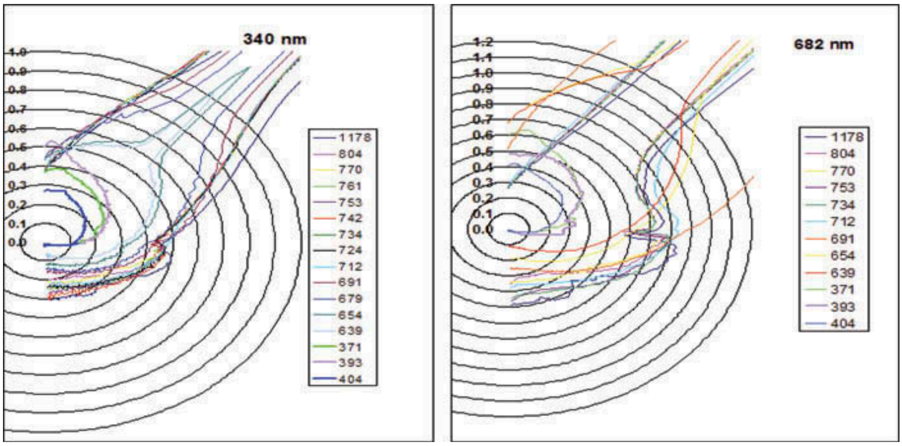

Figure 1. Transformation of diffuse radiance in channels 0.340 and 0.682 μm in cloud. Altitudes of observation in metres are in the legend.

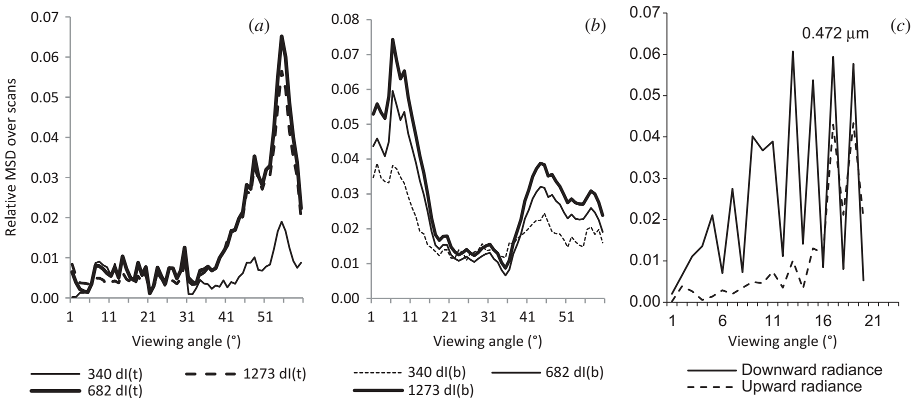

Figure 2. Relative errors of airborne measurement of diffuse solar radiance (*a*) above the cloud (reflected radiance at 799 m), (*b*) under the cloud (transmitted radiance at 370 m), and (*c*) within the cloud (upward and downward radiance at 670 m and in channel 0.472 μm). Spectral channels in (*a*) and (*b*) are indicated in the figure.

and 5 October 1972) (Kondratyev et al. 1977); Polar Experiment (POLEX) over the Kara Sea (75° N, 1 October 1972, 29 May 1977) above and under cloud layers (Grishechkin and Melnikova 1989); and experiments over Lake Ladoga (60° N, 24 September 1972 and 20 April 1985) (Kondratyev et al. 1977; Grishechkin, Melnikova, and Shults 1989). Observations over Lake Ladoga include measurements within clouds. Details of these experiments may be found in the reference material.

3. Observational uncertainties

A description of CAR is given by Gatebe et al. (2007). No dependency of the instrumental uncertainties on viewing angles under laboratory conditions was found, but during the flight the aircraft position was determined by roll, pitch, and yaw (Román et al. 2011; King 1987; Melnikova and Vasilyev 2004). Even with a stable flight, these angles are influenced by random variations. This introduced a random error of radiance measurements as a function of viewing directions. The possibility of estimating these errors is provided by several scans (4–10) registering at each altitude. Averaging the radiance for every viewing angle over all scans in a given altitude is accomplished together with the standard deviation, which is taken as the random measurement error. A similar procedure was carried out at each measurement level involved in the processing. Figure 2 shows the observational radiance errors *versus* the viewing angle for the reflected radiance above the cloud (marked 't'), transmitted radiance under the cloud (marked 'b'), and upward and downward radiance at 670 m within the cloud.

It should be pointed out that technical improvements over the last 30 years have been mainly concerned with miniaturization of electronics and the avoidance of mechanical components, but the optics and calibration procedures remain more or less unchanged. Thus the quality of the data obtained decades ago is defined by intelligent instrument construction, careful equipment preparation, calibration, and tuning. All these features were applied in the equipment used in the older Russian experiments. A detailed description of factors influencing observational errors and instrument calibration these experiments (Kondratyev et al. 1977; Grishechkin and Melnikova 1989; Grishechkin, Melnikova, and Shults 1989) is given by Melnikova and Vasilyev (2004). Airborne measurement errors vary over the spectrum and flight conditions. Here we only mention that they are within the range 1–3% in the visible and near IR and less than 5% in the UV range.

4. Method of processing experimental data

As mentioned above, most of the earlier retrieval approaches assumed significant limits for desired parameters: semi-infinite optical thickness, single scattering albedo equal to one unit (conservative scattering), absence of spectral dependence of the optical thickness, etc., and only one optical parameter could be retrieved (either optical thickness or single scattering albedo) (King 1987; King, Radke, and Hobbs 1990; Rozanov and Kokhanovsky 2004). Here the method described by Melnikova and Vasilyev (2004), Melnikova and Mikhailov (1994) and Melnikova, Domnin et al. (2000) is used for cloud optical parameter retrieval. It does not apply such restrictions, making it possible to acquire both optical parameters for every wavelength independently, and has already been applied to satellite data and ground-based observations of intensity and aircraft flux measurements of reflected and transmitted solar radiation. In this paper, to resolve the problem we used measurements from the top and below the cloud, as well as those within the cloud. Different expressions were derived for each case (Melnikova, Nikitin, and Gatebe 2009). The algorithm of the problem solution is presented by Genya, Melniova and Gatebe (2010) and Melnikova, Genya, and Gatebe (2011). Directly from the measurements, the following optical parameters are determined: $s^2 = (1–\omega_0)/[3(1–g)]$ = similarity parameter and $\tau' = 3(1–g)\ \tau_0$ = scaled optical thickness, where g is the phase function asymmetry parameter. Subject to the geometrical thickness Δz of the whole cloud or sub-layer thickness between measurements levels Δz_i, it is easy to derive volume absorption $\kappa = s^2\tau'/\Delta z$ and scattering coefficient $\alpha = \tau_0/\Delta z–\kappa$.

(1) *Observation above the cloud*. Multiangular data of reflected radiation at the cloud top give intensity, ρ_j in relative units of the incident solar flux in viewing angles $\theta_j = \arccos\mu_I$, the ratio of two differences $[\rho_\infty(\mu_1,\mu_0,\varphi) - \rho_1]/[\rho_\infty(\mu_2,\mu_0,\varphi) - \rho_2]$ is considered, where ρ_1 and ρ_2 are the observed intensities at two viewing angles $\arccos\mu_1$

arccosμ_2 and $\rho_\infty(\mu_j,\mu_0,\varphi)$ is the reflection coefficient of a semi-infinite conservative atmosphere ($\tau_0 = \infty$, $\alpha_0 = 1$), which can be calculated using approximation (Melnikova and Vasilyev 2004):

$$\begin{aligned}
\rho_\infty(\varphi,\mu,\mu_0) &= \rho_0(\varphi,\mu,\mu_0) - 4K_0(\mu)K_0(\mu_0)s + \frac{a_2(\mu)a_2(\mu_0)}{12q'}s^2,\\
\rho_0(\varphi,\mu,\mu_0) &= \rho^0(\mu,\mu_0) + \sum_{m=1}^{N}\rho^m(\mu,\mu_0)\cos m\varphi,\\
\rho^0(\mu,\mu_0) &= (\mu+\mu_0)^{-1}[(0.937\mu+0.529)(0.937\mu_0+0.529)\\
&\quad + g(1.19\mu\mu_0 - 0.74(\mu+\mu_0)+0.49)]\\
&\quad \text{for } \mu \geq 0.15 \quad \text{or} \quad \mu_0 \geq 0.15.
\end{aligned}$$

More details of this calculation are given by Melnikova, Dlugach, et al. (2000).

After algebraic transformations, expressions for the similarity parameter, s^2 (Melnikova, Domnin, et al. 2000) and scale optical thickness, τ' (King 1987) were obtained:

$$s^2 = \frac{[\rho_0(\varphi,\mu_1,\mu_0)-\rho_1]K_0(\mu_2) - [\rho_0(\varphi,\mu_2,\mu_0)-\rho_2]K_0(\mu_1)}{[\rho_0(\varphi,\mu_2,\mu_0)-\rho_2]K_0(\mu_1)\left(\frac{K_2(\mu_1)}{K_0(\mu_1)}-\frac{K_2(\mu_2)}{K_0(\mu_2)}\right) - \frac{a_2(\mu_0)}{12q'}[K_0(\mu_1)a_2(\mu_2)-K_0(\mu_2)a_2(\mu_1)]},$$

$$\tau' = (2s)^{-1}\ln\left\{\frac{m\bar{l}K(\mu_i)K(\mu_0)}{\rho_\infty(\varphi,\mu_i,\mu_0)-\rho_i} + l\bar{l}\right\}, \tag{1}$$

where q'' = 0.714; constants m and l are determined by the properties of the cloud being considered and are calculated by formulas (Melnikova and Vasilyev 2004) after determining parameter s, subscript j points that determination of τ' is possible for both viewing angles in the pair; $K(\mu)$, $K_0(\mu)$, and $K_2(\mu)$ – escape function and coefficients of its expansion over the small parameter s; $a_2(\mu)$ – the second coefficient in the expansion of the plane albedo function as follows:

$$\begin{aligned}
a_2(\mu) &= 3K_0(\mu)\left(\frac{3}{1+g}(1.271\mu - 0.9) + 4q'\right),\\
K_0(\mu) &= 0.797\mu + 0.442,\\
K_2(\mu) &= 5/3n_2(\mu^2+0.1).
\end{aligned} \tag{2}$$

The form of these functions is known and their values for fixed solar and viewing angles can be found in tables or calculated using the approximation (Melnikova and Vasilyev 2004).

(2) *Observation under the cloud*. Similar expressions for a case of multi-angular data of intensity σ_j of transmitted solar radiation in relative units of the incident solar flux (transmission coefficient) beneath the cloud were obtained by Melnikova, Domnin, et al. (2000):

$$s^2 = \left[\frac{\sigma_1 \bar{K}_0(\mu_2)}{\sigma_2 \bar{K}_0(\mu_1)} - 1\right] \frac{1}{\frac{\bar{K}_2(\mu_1)}{\bar{K}_0(\mu_1)} - \frac{\bar{K}_2(\mu_2)}{\bar{K}_0(\mu_2)}},$$

$$\tau' = s^{-1} \ln\left[\frac{\sqrt{4\sigma(\tau,\mu_j,\mu_0)^2 l\bar{l} + m^2 \bar{K}(\mu_j)^2 K(\mu_0)^2} + m\bar{K}(\mu_j)K(\mu_0)}{2\sigma(\tau,\mu_j,\mu_0) l\bar{l}}\right]. \tag{3}$$

The overbar in the function $\overline{K}(\mu)$ and the value $\bar{l}$ indicates the account of the ground albedo A.

It should be noted that a difficulty arises when the value of the surface albedo A is obtained from the measurements. The ground albedo is determined in terms of fluxes – in our case, observations include the radiation intensity. However, the relation $A = I(48°)/I(132°)$, which is used here to determine the surface albedo from observations, was obtained earlier (Melnikova and Vasilyev 2004; Varotsos et al. 2013). It uses observation obtained beneath the cloud at two viewing angles, 48° and 132°. Mean albedo is calculated for all scans at a selected altitude of 370 m, the mean square deviation (MSD) of the surface albedo is estimated, and both parameters are presented in Table 1 in different spectral channels. It is seen that the standard deviations are appropriately low.

(3) *Observation inside the cloud.* Below are presented the formulas for determining the optical parameters from measurements inside the cloud. Expressions for parameter retrieval are different for the cloud sub-layer adjacent to the top, base, and remote from boarders. Denoting $I_i^{\downarrow} = I(\mu)$, $I_i^{\uparrow} = I(-\mu)$ and $J_i = I_i^{\downarrow} - I_i^{\uparrow}$. Here the number of cloud sub-layers between observational levels is n, and subscript i indicates the number of observational levels within the cloud.

For the upper sub-layer adjacent to the cloud top (Melnikova, Nikitin, and Gatebe 2009):

$$s^2 = \frac{9\mu^2(\rho_0 - \rho)^2 - 4J_1^2 K_0^2(\mu)}{9\mu^2\left\{16K_0^2(\mu)\left(K_0^2(\mu_0) - 0.25(I_1^{\uparrow} + I_1^{\downarrow})^2\right) - 2\frac{K_2(\mu)}{K_0(\mu)}(\rho_0 - \rho)^2 - \frac{a_2(\mu_0)a_2(\mu)}{6q'}(\rho_0 - \rho)\right\}},$$

$$\tau'_1 = \frac{1}{2s} \ln\left[l \frac{\left[I_1^{\downarrow}(1 - 6\mu s + 18\mu^2 s^2) - I_1^{\uparrow}\right]}{\left[I_1^{\downarrow} - (1 - 6\mu s + 18\mu^2 s^2) I_1^{\uparrow}\right]} \frac{(\rho^{\infty} - \rho) + 4K_0(\mu_0)K_0(\mu)s + \frac{a_2(\mu_0)a_2(\mu)}{12q'}s^2}{(\rho^{\infty} - \rho) - 4K_0(\mu_0)K_0(\mu)s + \frac{a_2(\mu_0)a_2(\mu)}{12q'}s^2}\right]. \tag{4}$$

Table 1. Atlantic sea surface albedo obtained from CAR observations.

λ (µm)	**0.340**	**0.381**	**0.472**	**0.682**	**0.870**	**1.035**	**1.219**	**1.273**
A_s	0.0662	0.0673	0.0505	0.0458	0.0430	0.0417	0.0403	0.0435
MSD	0.0036	0.015	0.0012	0.0012	0.0012	0.0012	0.0010	0.0012

For the internal cloud sub-layers:

$$s^2 = \frac{\left(J_i^2 - J_{i-1}^2\right) J_i^2 J_{i-1}^2}{36\mu^2 \left[J_{i-1}^2 I_i^{\downarrow} I_i^{\uparrow} - J_i^2 I_{i-1}^{\downarrow} I_{i-1}^{\uparrow}\right]},$$

$$\tau_i' - \tau_{i-1}' = \frac{1}{2s}\ln\left[\frac{\left[J_{i-1} - 6I_{i-1}^{\downarrow}\mu s(1-3\mu s)\right]}{\left[J_{i-1} + 6I_{i-1}^{\uparrow}\mu s(1-3\mu s)\right]}\right] + \frac{1}{2s}\ln\left[\frac{\left[J_i + 6I_i^{\uparrow}\mu s(1-3\mu s)\right]}{\left[J_i - 6I_i^{\downarrow}\mu s(1-3\mu s)\right]}\right]. \tag{5}$$

For the lower sub-layer adjacent to the cloud base:

$$s^2 = \frac{\mu^2\sigma^2 - \frac{4}{9}\bar{K}_0^2(\mu) J_{n-1}^2}{\frac{2\mu^2\sigma^2}{\bar{K}_0(\mu)}\left[K_2(\mu) + A\frac{a_2(\mu)+n_2+24\bar{K}_0(\mu)\bar{q}'}{1-A} + \frac{12q'A^2}{[1-A]^2}\right] - 9q'\mu^2\sigma^2 - 4\mu^2 J_{n-1}^2\left(I_{n-1}^{\downarrow} + I_{n-1}^{\uparrow}\right)^2},$$

$$\begin{aligned}\tau_n' - \tau_{n-1}' = {} & \frac{1}{2s}\ln\left[\frac{2}{3\mu\sigma}\frac{I_{n-1}^{\downarrow} - I_{n-1}^{\uparrow}(1-6\mu s+18\mu^2 s^2)}{1-3\mu s+9\mu^2 s^2}\right] \\ & + \ln\Bigg[K_0(\mu)(1-3q's) + K_2(\mu)s^2 + \frac{A}{1-A}\left(1-3q's+n_2 s^2\right) \\ & (1-4K_0(\mu)s + a_2(\mu)s^2)\Bigg].\end{aligned}$$

Observations at every observational level within the cloud are used for retrieving optical parameters of the layer between observational levels for all chosen viewing angles. Then the values obtained are averaged taking into account observational and retrieval errors.

5. Accounting for cloud layer horizontal heterogeneity

The horizontal heterogeneity of the cloud top increases the proportion of diffuse radiation r illuminating the cloud, which should be considered in the calculation of the functions determining the geometry of the problem $K(\mu)$, $K(\mu_0)$, $a_2(\mu)$, $a_2(\mu_0)$, $\rho_0(\mu,\mu_0)$ (Melnikova and Vasilyev 2004; Melnikova, Dlugach, et al. 2000):

$$\begin{aligned}\rho(\mu,\mu_0) &= \rho(\mu,\mu_0)(1-r) + ra(\mu_0), \\ K(\mu_0) &= K(\mu_0)(1-r) + rn, \\ a(\mu_0) &= a(\mu_0)(1-r) + ra^{\infty},\end{aligned} \tag{7}$$

where spherical albedo a^{∞}, plane albedo $a(\mu_0)$, and the value n are determined as cosine μ weight-average of the viewing angle of the corresponding functions:

$$a^{\infty} = 2\int_0^1 a(\mu_0)\mu_0 d\mu_0 = 4\int_0^1 \mu_0 d\mu_0 \int_0^1 \rho(\mu,\mu_0)\mu d\mu\,,\ n = 2\int_0^1 K(\mu)\mu d\mu.$$

When this approach applied to observations of the reflected radiation it corresponds to cloud top boundary heterogeneity. When the approach is applied to the data of transmitted radiation it characterizes cloud thickness heterogeneity.

To determine the shadow parameter of cloud horizontal heterogeneity, r, optical thickness is obtained from data of reflected radiance (the shadow parameter of cloud thickness heterogeneity is determined from observations of transmitted radiance) on the assumption of conservative scattering. The shadow parameter is given by

$$r = \frac{1}{N\bar{\tau}} \sqrt{\sum_{i=1}^{N} |\tau_i - \bar{\tau}|^2}, \tag{8}$$

where N is the number of considered viewing directions, and optical thickness is retrieved for every viewing direction by the assumption of the conservative scattering in the first step of processing. The mean value of the conservative optical thickness and parameter r with respect to the viewing angles is shown in Table 2. Obtained values show a weak heterogeneity as of whole cloud from observations under the cloud ($r < 0.01$) as of cloud top ($r \sim 0.02$) from observations above the cloud. Taking into account the cloud heterogeneity alters retrieved optical thickness values by 10% on average.

6. Regularization of solution

As mentioned above, random variations in the angles that define a plane's flight affect measurement uncertainty and provoke the angular dependence of errors of observed intensity. The estimation of these errors is provided by registration of several scans at every observational level. Figure 2 shows the MSD of the intensity, obtained by averaging over scans measured at an altitude of 800 m above the cloud (Figure 2(*a*); denoted (t)), at 370 m under the cloud (Figure 2(*b*); denoted (b)), and at 0.670 km within the cloud (Figure 2(*c*)) in the 0.472 μm channel. In Figures 2(*a*) and (*b*) spectral channels are indicated in the legend. The observational errors were used in the regularization procedure together with the retrieval errors. This involves the use of the following formula (Vasilyev and Melnikova 2009):

$$x = \frac{\sum_j x_j \frac{1}{(\Delta x_j)^2}}{\sum_j \frac{1}{(\Delta x_j)^2}}, \tag{9}$$

where x_j is the retrieved parameter for the jth viewing angle (or jth pair of angles) and Δx_j is the sum of the errors of measurements and processing. These are calculated by expressions for indirect errors, obtained from Equations (1)–(6) for retrieving the required parameters (τ' and s^2). Analytical presentation of the inversion provides formulas for the errors caused by the proposed procedure of data processing (Melnikova and Vasilyev 2004; Melnikova, Nikitin, and Gatebe 2009). For the uncertainty of parameter s^2, the case of observation under the cloud gives the following:

$$\frac{\Delta s}{s} = \frac{\Delta\sigma \bar{K}_0(\mu_1) + \sigma\Delta\bar{K}_0}{\sigma_1 \bar{K}_0(\mu_2) - \sigma_2 \bar{K}_0(\mu_1)} + \frac{\Delta\sigma}{2\sigma} + \frac{\Delta\bar{K}_0}{2\bar{K}_0(\mu)} + \frac{2\mu\Delta\mu}{\left(\frac{\bar{K}_2(\mu_1)}{\bar{K}_0(\mu_1)} - \frac{\bar{K}_2(\mu_2)}{\bar{K}_0(\mu_2)}\right)^2} \tag{10}$$

and for optical thickness:

Table 2. Cloud optical parameters retrieved from radiative airborne observations above and under the cloud.

λ (μm)		0.340	0.381	0.472	0.682	0.870	1.035	1.219	1.273
$\tau_{Conserv}$	Base	25.9	17.6	15.8	11.9	10.2	10.5	12.4	13.6
	Top	5.1	7.18	12.7	12.4	12.9	16.5	10.7	10.3
τ'	Base	45.2	36.1	35.6	30.4	26.2	28.1	27.6	29.7
	Top	6.6	7.4	10.8	13.6	13.1	16.8	11.4	10.7
$D\tau'/\tau$		0.010	0.006	0.005	0.004	0.003	0.004	0.005	0.004
σ (km^{-1})	Base	113.05	90.27	88.84	76.11	65.63	70.16	69.00	74.18
	Top	16.63	18.58	27.12	34.00	32.80	41.88	28.42	26.77
r	Base	0.0074	0.010	0.010	0.009	0.008	0.013	0.014	0.012
	Top	0.023	0.012	0.010	0.018	0.023	0.008	0.012	0.009
ω_0	Base	0.99931	0.99896	0.99917	0.99871	0.99904	0.99831	0.99831	0.99859
	Top	0.99177	0.99911	0.99850	0.99911	0.99961	0.99988	0.99905	0.99906
$1-\omega_0$	Base	0.00069	0.00104	0.00083	0.00129	0.00096	0.00169	0.00169	0.00141
	Top	0.00823	0.00089	0.00015	0.00089	0.00039	0.00012	0.00095	0.00094
κ (km^{-1})	Base	0.04672	0.07234	0.06217	0.10697	0.07491	0.13547	0.13716	0.12437
	Top	0.16601	0.01629	0.01698	0.01438	0.01632	0.015761	0.02854	0.03462
$D\omega/\omega$		0.0002	0.0004	0.0002	0.0004	0.0002	0.0004	0.001	0.0007

$$\frac{\Delta\tau_0}{\tau_0} = \frac{\Delta s}{s} + \frac{1}{\tau_0}\left\{2\frac{\Delta\sigma}{\sigma} + 8\frac{\Delta s}{s} + \frac{\Delta\bar{K}}{\bar{K}(\mu)} + \frac{\Delta K}{K(\mu)}\left[\frac{\bar{l}l\left(\frac{2\sigma}{m\bar{K}(\mu_j)K(\mu_0)}\right)^2}{\left(\sqrt{\bar{l}l\left(\frac{2\sigma}{m\bar{K}(\mu_j)K(\mu_0)}\right)^2+1}+\bar{l}l\left(\frac{2\sigma}{m\bar{K}(\mu_j)K(\mu_0)}\right)^2+1\right)}+1\right]\right\} \tag{11}$$

$\Delta\sigma/\sigma$ is the observational uncertainty. The analogous expressions are derived for all other cases of observation used for retrieval. Uncertainties calculated by these formulas are used for the regularization of the results in the program algorithm.

Figure 3 shows non-regularized results of retrieved optical thickness and parameter s^2 *versus* viewing angles. It is clear that only the optical thickness obtained from measurements under the cloud (Figure 3*a*, base) does not depend on the viewing angles (as it should be, ideally). The regularization in this example gives the value of parameter s^2 = 0.001568 (base) and 0.001469 (top), and simple averaging without accounting for the errors' dependence on the viewing angles, brings the value to 0.002004 (base) and 0.001327 (top). The differences are 22% and 10%, respectively. Here we place no restriction on the parameter s^2 (even the demand of a positive value).

7. Results

The three parameters (ground albedo A, similarity parameter s^2, and scaled optical thickness τ') for the whole cloud are obtained with the use of the above formulas from observations at the cloud top (T) at altitude 800 m, and at the cloud base (B) at 370 m. This approach allows all suitable pairs of intensities of radiation in two viewing angles in Equations (1) and (3) to be successively sampled. Measurement data are analysed in the range of zenith (for transmitted radiation) and nadir (for reflected

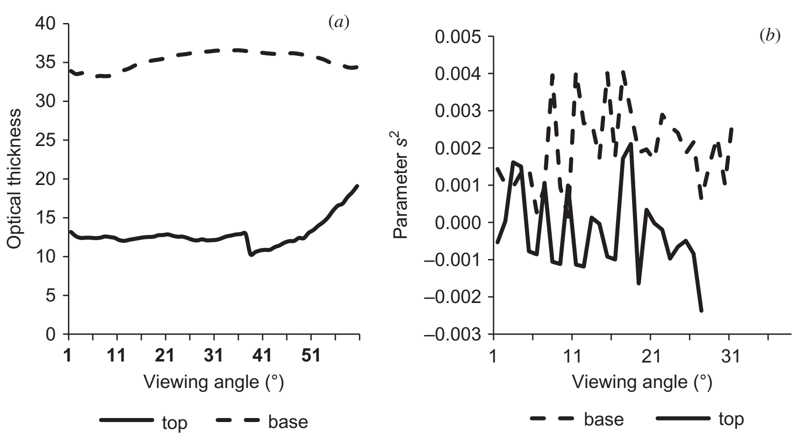

Figure 3. Optical parameters retrieved for different viewing angles in the channel 472 nm: top – obtained from observation above the cloud; base – from observations below the cloud: (*a*) optical thickness, (*b*) the similarity parameter s^2.

radiation) viewing angles from 0 to 70º. Viewing angles very close to the horizon were not considered due to increased uncertainties in the plane layer model. During processing, we consider only those pairs of viewing angles that meet the proximity values of the optical thickness determined in the approximation of conservative scattering (to 0.5%) and the difference of cosines of viewing angles in a pair more than 0.1. Next, the obtained values of the required parameters are averaged over all considered pairs with a weight equal to the inverse value of the error for the solution regularization. The processing algorithm provides assessment of the standard deviations of the results. The results are presented in Table 2 and Figure 4.

To obtain the single scattering albedo α_0, the optical thickness τ, and the scattering coefficient α, the spectral values of the asymmetry parameter g were taken into account according to Lobanova et al. (2009) and Stephens (1979).

Figure 4(*a*) presents the results obtained from observations by NASA (triangles) and the spectral dependence of the optical thickness derived from the Russian airborne data (geographical sites and dates are in the figure).

Figure 5(*a*) demonstrates spectral values of the water surface albedo obtained by the procedure described above and older Russian albedo spectral dependences directly calculated from the ratio of upward and downward fluxes at the under-cloud observational level (300–500 m, indicated in figure legend) (Varotsos et al. 2013) and albedo retrieved from radiance NASA measurements as described above.

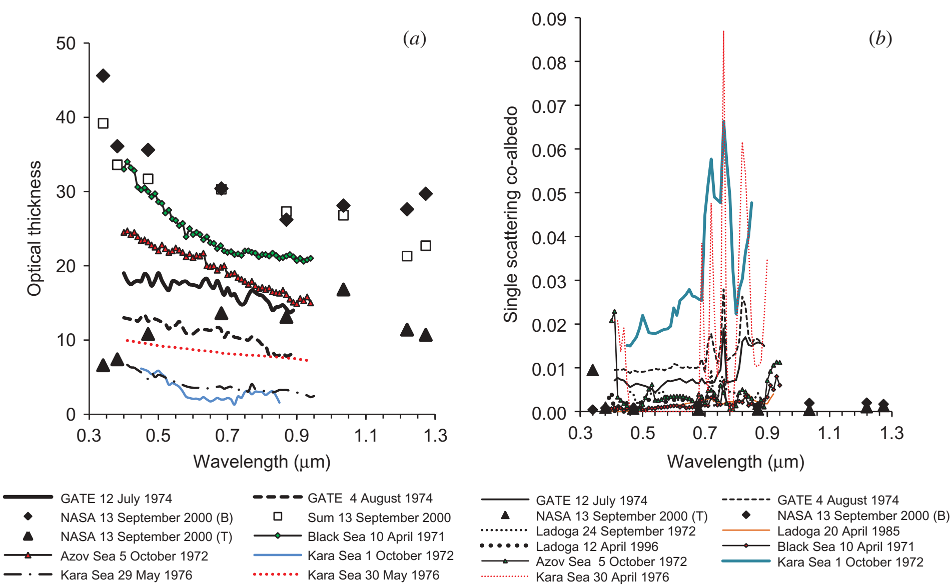

Figure 4. Spectral values of (*a*) optical thickness and (*b*) single scattering co-albedo ($1–\alpha_0$) for cloud, retrieved from airborne observations close to the south-west African coast on 13 September 2000 from base (B) and top (T) observation, the sum of optical thickness over all sub-layers within the cloud (Sum), and spectral dependence of optical parameters obtained from older Russian experiments above the Atlantic Ocean close to northwest Africa, 1974, the Black, Azov, and Kara Seas 1971, 1972, and 1976, and Ladoga Lake, 1972 and 1985.

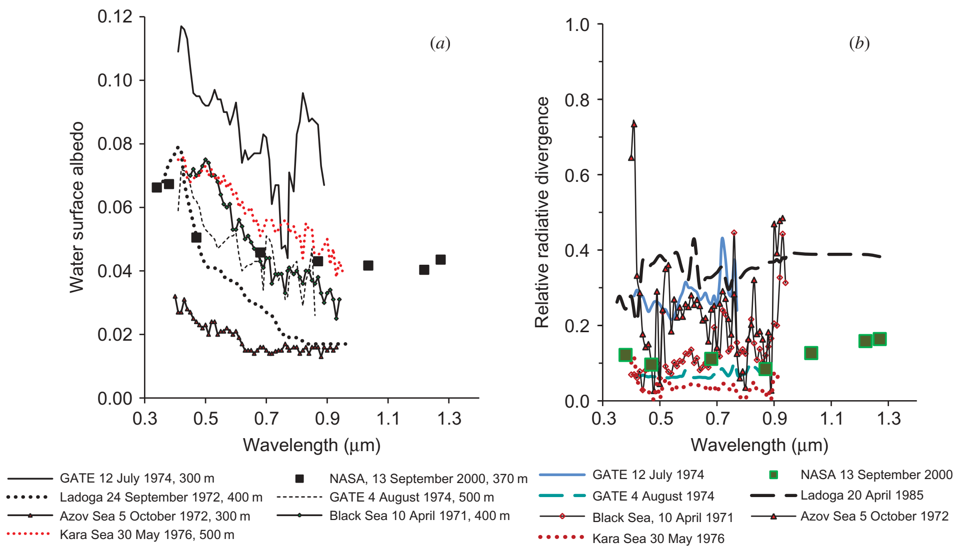

Figure 5. Spectral dependence of (*a*) water surface albedo and (*b*) relative radiative divergence from airborne experiments indicated in figure.

8. Vertical profile of cloud optical parameters

As mentioned above, measurements within the cloud were obtained too close to the top (altitude 0.8–0.6 km) and do not satisfy the diffuse domain condition. Nevertheless, we attempted to retrieve the optical parameters from measurements within the cloud for viewing directions close to nadir and zenith, which turned out to be successful for most sub-layers between the measurement levels. The averaging and estimation of the standard deviation is accomplished over six scans at every level. The special procedure, analogous to regularization while taking into account average error at every height, is applied to vertical profiles for smoothing. Figures 6(*a*) and (*b*) show vertical profiles of the volume absorption and scattering coefficients retrieved from NASA observations on 13 September 2000.

The scattering coefficient is sufficiently high in the altitude range 0.685–0.750 km. Variations are similar at all spectral channels that validate each other and may indicate cloud vertical heterogeneity. The scattering coefficient values 300–500 km^{-1} are reliable for tropical cloud (droplet sixe is about 10 µm, concentration about 500 cm^{-3}, providing a scattering coefficient of about 300 km^{-1}).

There are two maximum absorption coefficients at altitudes 0.647 and 0.685 km, the lower one coinciding with the low maximum of the scattering coefficient. It is necessary to take into account the fact that observations taken too close to the cloud top and breaking the asymptotic formulas applicability region (diffuse domain) increase retrieval errors. The experience of retrieval indicates that no positive continuous solution exists if the applicability region breaks out. Here we have a result, but with increased uncertainty. It is important that the approach appears effective even for under less than perfect conditions and could be recommended for other similar observational data processing.

Obtained profiles are compared to optical parameters retrieved from airborne spectral irradiance observations above Lake Ladoga on 24 September 1972 and 20 April 1985 in the range 0.330–0.960 µm, as shown in Figures 6(*c*) and (*d*) (Melnikova and Vasilyev

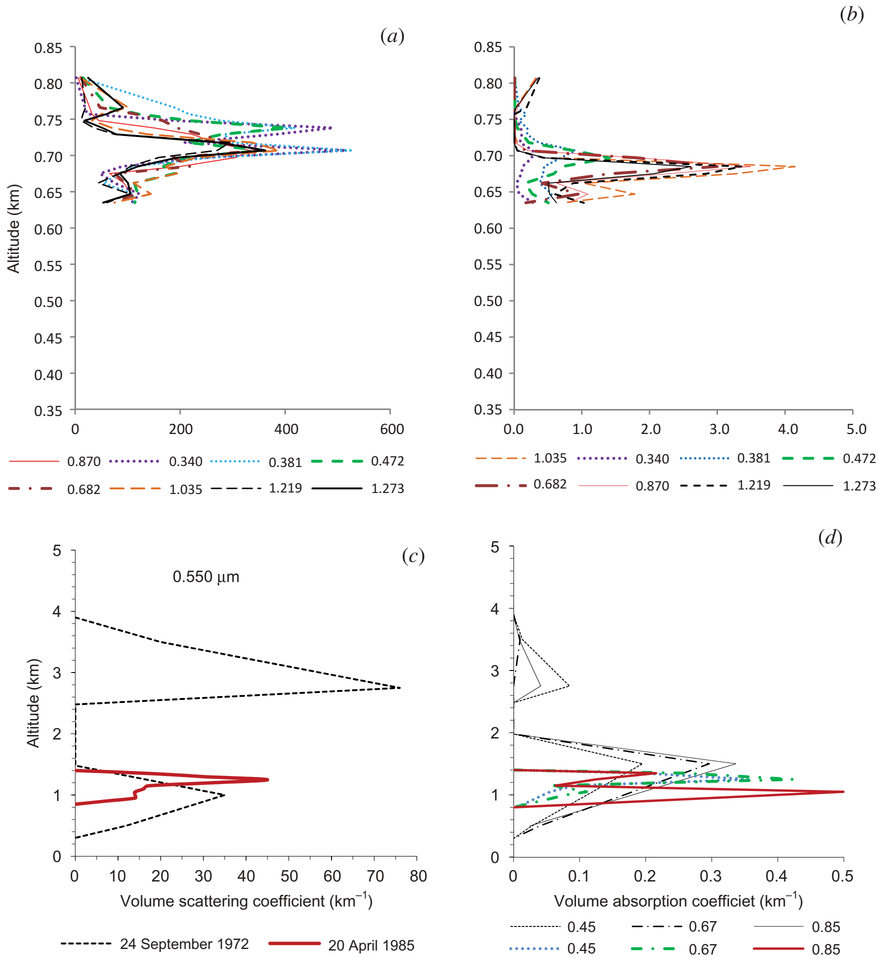

Figure 6. Vertical profile of the volume scattering coefficient (*a, c*) and volume absorption coefficient (*b, d*) from airborne observation close to the southwest African coast on 13 September 2000 (*a, b*) and over Ladoga Lake on 24 September 1972 (thin lines) and 20 April 1985 (*c, d*) (thick lines). Legend in (*a*), (*b*), and (*d*) indicates wavelength.

2004; Melnikova and Mikhailov 2001). It is seen that volume scattering and absorption coefficients in tropical regions are an order of magnitude higher than at high latitudes. We draw attention to the triangular shape of the vertical profiles, which is similar to that obtained in high-level clouds in Feofilov et al. (2013).

9. Analysis of optical cloud parameters retrieved

Figure 4(*a*) presents the optical thicknesses obtained from different experimental data. The difference in values appears in different geographical latitudes. Most of the data show a clear spectral dependence on optical thickness that agrees with results obtained from data from other instruments and processing methods (Melnikova and Vasilyev 2004; Melnikova and Mikhailov 1994; Grishechkin and Melnikova 1989; Grishechkin, Melnikova, and Shults 1989). The summation of the optical thicknesses of sub-layers between observational levels retrieved from NASA observations within cloud gives the values denoted in Figure 4(*a*) by white squares, for additional validation. Optical thickness obtained from base and summation from inside observations practically coincide, but results retrieved from top data from NASA observations on 13 September 2013 appear essentially lower, especially in the UV and visible spectral ranges, which on the one hand, reflects the vertical heterogeneity of the cloud layer, but on the other hand, reflects the effect of measurement errors. Above the cloud, there is a greater influence of inhomogeneity on the upper boundary of the cloud layer, and higher turbulence affects aircraft below the cloud. Possibly this expresses the horizontal inhomogeneity of the cloud because the time difference between measurement sites at the top and base is 20 min and the distance is 80–90 km.

Figure 4(*b*) shows values of the single scattering co-albedo obtained from top and base NASA observations, which appear very close and might be considered as mutual validation. The spectral values of the single scattering co-albedo ($1–\alpha_0$) obtained from the data of Russian experiments in the tropics are twofold greater. It should be mentioned that observations in the GATE programme were carried out after a sandstorm in the Sahara (Kondratyev et al. 1977), so the absorption in cloud is greater than in the case of observations by NASA in 2000. Clouds in experiments over the Kara Sea in 1972 and 1976 are characterized by a low scattering coefficient (Figure 7*b*) that yields increased ($1–\alpha_0$) values (blue and dotted red lines).

Figures 7(*a*) and (*b*) demonstrate spectral dependence of the volume absorption and scattering coefficients obtained from the above-mentioned airborne experiments. Solid

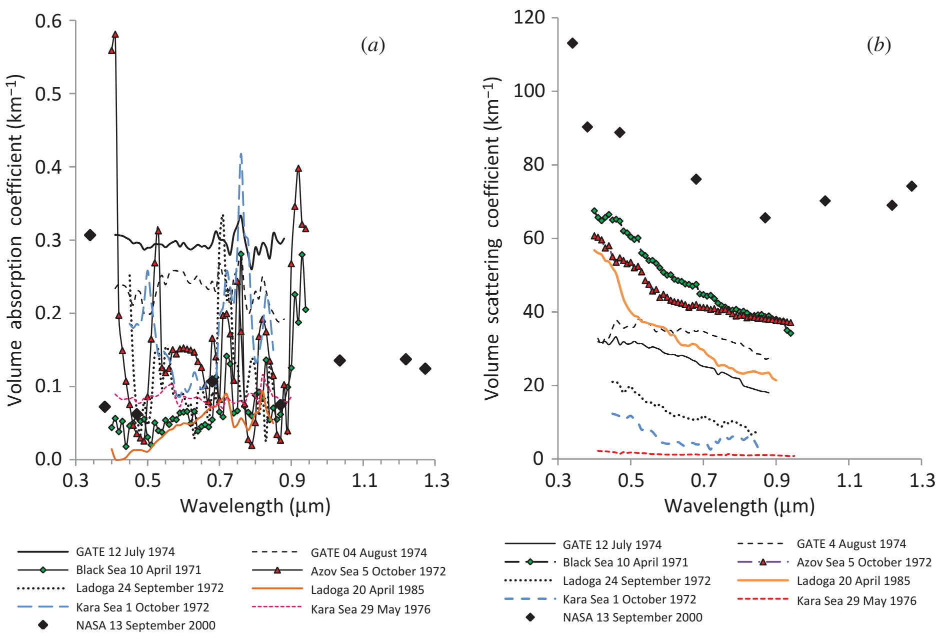

Figure 7. Spectral dependence (*a*) volume absorption coefficient, (*b*) volume scattering coefficient of extended clouds retrieved from airborne observation close to the west coast of Africa (northwest, GATE, 1974) and (southwest, NASA, 2000), above the Black, Azov, and Kara Seas, 1971, 1972, and 1976, the Ladoga Lake, 1972 and 1985.

rhombs correspond to results retrieved from the radiance data of NASA on 13 September 2000, while curves indicate results retrieved from spectral irradiance data of older Russian experiments that are indicated in the legend. The absorption coefficient of molecular absorption bands is less than 0.1 km^{-1} for all results apart from GATE, 1974. It seems that only an increased content of dust aerosols in the atmosphere during GATE 1974 increases the absorption coefficient twofold compared with other results. The ozone absorption manifests itself at curves corresponding to higher clouds (the top higher than 0.8 km) at wavelength <0.4 μm (appears also in the experiment of NASA on 13 September 2000) and at 0.6–0.65 μm. The uptake of the ozone into aerosol and water droplets demonstrates the significance of ozone–cloud climate interactions in accordance with Kondratyev and Varotsos (1996). The Chappuis band of ozone (O_3) is distinguished at wavelength $\lambda \sim 0.60$–0.65 μm in curves corresponding to experiments GATE, 4 August 1974, the Azov Sea, 5 October 1972, the Black Sea 10 April 1971, and Lake Ladoga, 20 April 1985, that indicate tropospheric ozone influence (Varotsos 2005). This might explain the fact that in the presence of thick clouds the column ozone content deduced from satellite observations is generally underestimated. This underestimation of total ozone was quantitatively examined by Varotsos (1995), especially in synoptic cases where ozone-rich air has been transported into the lower troposphere.

The relative radiative divergence obtained directly from observed irradiance at the cloud top and base and calculated with the Delta-Eddington method from retrieved cloud optical parameters is shown in Figure 5(*b*). It is seen that a significant radiation absorption in cloud layers confirms the underestimation of cloud–aerosol–radiation interaction in climate models (Kondratyev and Varotsos 1995).

The scattering coefficient is sufficiently low (<5 km^{-1}) at high latitudes, increases at middle latitudes (~12 km^{-1}) and in subtropical regions (~20 km^{-1}), and reaches 70 km^{-1} in the tropics – possibly explained by atmospheric moisture. Certainly, more statistics are needed for a reliable conclusion, because water vapour content is influenced by many geographical factors. All results demonstrate the transparent spectral dependence of the scattering coefficient.

10. Conclusion

These airborne data allow a comprehensive cloud analysis in terms of a rigorous radiation transfer theory. The experiments were the basis for creating a detailed data processing algorithm of similar measurements. Note that the intensity measurement is more strongly influenced by random variations in the orientation of the aircraft than the measurement of hemispherical flow. However, the registration of several scans at each height to evaluate the emerging random error, and a large number of viewing directions, allows the regularization of the results. The data processing procedure is determined by measurement errors and methods for solving the inverse problem.

Note that the analytical approach for inverting observational data (obtaining the optical parameters of clouds and their vertical profile) has advantages over the methods used by other authors. In particular, it does not have additional restrictions and links to required parameters, thus providing results closer to real nature. The formulas for uncertainties are directly obtained from the formulas for determining the desired parameters, which allow transparent analysis of the solution's stability and its regularization.

The cloud optical parameters obtained in the application of the analytical inverting method to NASA measurements agree satisfactorily with previously obtained values from the data of spectral semispherical flux measurements by other instruments and in other experiments. The comparison of volume absorption coefficient shows the uniformity of values in different regions. It is revealed from analysis of results in different regions (northwest and southwest Africa, North Europe) that the volume scattering coefficient increases from polar to tropical regions. The compatibility (not coincidence) of the results obtained in different seasons (April 1971, September and October 1972, July and August 1974, April 1985, and September 2000) might point to a weak temporal variability of the main parameters of stratus clouds.

Retrieved values of the single scattering albedo and absorption coefficients in most experiments provoke a significant solar radiation absorption in cloudy atmosphere, which needs detailed study at the global scale and accurate statistical processing (e.g. in accordance with Prasolov and Khovanov (2008) and Prasolov and Wei 2000)).

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